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Impact of the lateral blast on the spatial pattern and grain size characteristics of the May 18, 1980 Mount St. Helens fallout deposit

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ABSTRACT

The May 18, 1980 eruption of Mount St. Helens started with a lateral blast that fed a pyroclastic surge, which then uplifted to form a co-blast plume. 30 minutes later, Plinian activity started at the vent and fluctuated in intensity for ~ 9 hours. The resulting fallout deposit, documented to > 600 km from vent, presents some striking features: 1) displacement of the overall deposit to the north of the vent; 2) a secondary thickness and mass maximum at ~300 km from vent; 3) a total grain-size distribution dominated by fine ash (62 wt% of the deposit < 63 μm); and 4) individual grain-size distributions that vary dramatically in the cross-wind direction from strongly bimodal in the south to skewed unimodal in the north. Results from a new deconvolution of the individual grain-size distributions show that they are a combination of a coarse subpopulation that decreases in size with distance from vent and a constant fine subpopulation with a mean ~ 15 μm . Relative proportions of each subpopulation vary asymmetrically in the cross-wind directions, with the fine subpopulation preponderant towards the north and the coarse one dominating the south of the deposit; both reach their absolute maxima in mass on the deposit axis. Componentry analyses of selected samples show that blast-derived material is greatly enriched towards the north of the deposit. These results indicate that the co-blast plume dispersed fine-grained material over great distances and dominated the fine subpopulation. Comparison with reanalysis data of atmospheric wind fields and satellite images of the spreading ash cloud suggests contrasting ash transport and depositional processes for the (early) co-blast plume and the (later) vent-derived Plinian plumes. The co-blast plume is displaced to the north; it had a high overshoot height, and eastward dispersion via strong winds low in the stratosphere (~10-15 km). The Plinian plumes were lower and dispersed most of the material to the southeast as the direction of high velocity winds shifted just before the late climactic Plinian eruptive phase. Fine ash (fine subpopulation) was deposited continuously throughout the deposit, with an increase of sedimentation rate ~ 300 km from the vent where there is a secondary maximum in the deposit mass and thickness. Fine ash probably settled by a combination of enhanced sedimentation mechanisms, including not only aggregation but also gravitational convective instabilities of cloud base, hydrometeor formation and destruction, and entrainment of small particles by larger ones. Finally, we show that half of the deposit (by mass) in the medial area was deposited by the co-blast plume, and that a significant proportion of the Mount St. Helens fallout deposit is non-juvenile, which has implications for the magmatic budget of this eruption.

1 Introduction

The May 18, 1980 eruption of Mount St. Helens (USA) was a key event for the development of modern volcanology and has produced major scientific advances regarding the dynamics of explosive eruptions. In particular, many concepts underpinning our understanding of ash transport and deposition from volcanic plumes stem from insights gained from the extraordinary quality of the field observations, including both field measurements and remote sensing data collected during and after the event. One example is the commonly accepted theory that distal mass accumulation maxima in tephra fall deposits are the signature of premature fallout of fine material due to aggregation (Brown et al., 2012; Durant and Rose, 2009), which was first developed to explain the depositional pattern of the May 18, 1980 Mount St. Helens fallout (e.g., Carey and Sigurdsson, 1982). Also ground-breaking was the exceptional characterization of ash transport : the May 18, 1980 Mount St. Helens eruption cloud was one of the first to be observed by airborne remote sensing methods (i.e. radar and satellite imagery; Harris et al., 1981; Holasek and Self, 1995; Rice, 1981; Sparks et al., 1986) and sampled in-situ by research aircraft (Hobbs et al., 1981; Hobbs et al., 1982; Farlow et al., 1981). The measured airborne grain-size distribution has become one of the references for the grain-size distribution input for the London Volcanic Ash Advisory Centre (VAAC) ash dispersion model used operationally today (Beckett et al., 2014). Finally, the tephra fall deposit is one of the most extensively mapped and sampled modern fallout deposits, with quantitative field data up to distances > 700 km from vent (Carey et al., 1990; Carey and Sigurdsson, 1982; Sarna-Wojcicki et al., 1981). As a result of the quality of both airborne and ground observations, this eruption has been used to benchmark several generations of numerical models of ash dispersion and sedimentation (Armienti et al., 1988; Fero et al., 2008; Folch et al., 2010; Mastin et al., 2009); the total grain-size distribution is also used as input for simulations of much larger eruptions (Mastin et al., 2014).

Accurate forecasts of atmospheric ash quantities are required during volcanic crises to safely manage flight operations because of the recognized impact of ash ingestion into airplane engines (Casadevall et al., 1996; Gislason et al., 2011; Kueppers et al., 2014). Aircraft operators require not only predictions of ash cloud transport but, most importantly, ash concentration to anticipate potential airspace closures (Beckett et al., 2014; Witham et al., 2007). One consequence of these new requirements is the need for accurate input data to initialize both operational numerical models for ash dispersion (Mastin et al., 2009; Stohl et al., 2011) and algorithms for ash retrieval from satellite images (Guéhenneux et al., 2015; Labazuy et al., 2012; Mackie and Watson, 2014). Discrepancies between predictions of ash dispersion (spatial

distribution, concentration and grain size) and ash measured and sampled on the ground during recent volcanic crises (Stevenson et al., 2015), have highlighted the limitations of the commonly used eruption source parameters. Further problems come from modelling the sink term used to forecast volcanic ash cloud transport, which is not accurately represented in operational models and may thus overestimate far-field airborne ash quantities (Durant, 2015). Especially critical are (1) the initial grain-size distribution and (2) the mass eruption rate. The former is generally approximated using total grain-size distributions from known eruptions (for proximal dispersion) (Bonadonna et al., 2015; Bonadonna and Houghton, 2005) and airborne (in situ) measurements of grain sizes (for distal clouds) (Flentje et al., 2010; Hobbs et al., 1981; Hobbs et al., 1982; Johnson et al., 2012). The latter is derived from the column height by an empirical law with a theoretical basis (Degruyter and Bonadonna, 2012; Mastin et al., 2009; Sparks et al., 1997; Woodhouse et al. 2013). Yet, recent work has shown that both the dispersion of volcanic eruption clouds and sedimentation of ash from those clouds are spatially and temporally complex because of gravitational instabilities of the cloud base (Carazzo and Jellinek, 2012; Carazzo and Jellinek, 2013; Chakraborty et al., 2006; Manzella et al., 2015), ice nucleation on ash particles and formation of hydrometeors (Bingemer et al., 2012; Durant, 2015; Durant et al., 2009; Durant et al., 2008), and the variety and time-dependence of aggregation processes (Brown et al., 2012). Furthermore, the addition of fine ash transported in plumes uplifting from pyroclastic density currents (PDCs) during their propagation (Andrews and Manga, 2012; Bonadonna et al., 2002; Sparks and Walker, 1977), hereafter generically referred to as co-PDC plumes, will affect the grain size, mass and energetics of the dispersing plume, and yet has rarely been accounted for in plume models.

In light of these new findings, and to investigate the role and impact of co-PDC plumes on ash transport during a well-documented modern eruption, we return to the iconic eruption of Mount St. Helens to re-examine and decipher its distal fallout deposit. Despite being a ‘textbook’ case of a Plinian eruption, the May 18, 1980 eruption of Mount St. Helens had unusual dynamics and produced a complex fallout deposit. In addition to its remarkable secondary thickness maximum ~300 km from the vent, the fallout deposit presents several striking features. First, the overall deposit is displaced to the north of the vent (Carey and Sigurdsson, 1982; Sarna-Wojcicki et al., 1981), a feature that has yet to be included in ash dispersion models except by moving the presumed vent for the entire eruption to the north (Armienti et al., 1988; Fero et al., 2008; Folch et al., 2010). Second, the total grain-size distribution, which contains 62% by mass of fine ash ($< 63 \mu\text{m}$), makes it one of the most fines-rich fallout deposits ever characterized (Rust and Cashman, 2011). Finally, individual grain-

size distributions vary dramatically not only downwind, but also in the cross-wind direction (Durant et al., 2009). These features have never been satisfactorily reproduced by numerical models of plume dispersion (Armienti et al., 1988; Fero et al., 2008; Folch et al., 2010) except by using effective grain sizes (and settling velocities) back-calculated from the deposit data (Hopkins and Bridgman, 1985).

To re-examine key features of the deposit, we use a new method to deconvolve individual grain-size distributions and combine these data with componentry analyses of selected cross- and down-wind samples. We then compare the field observations with reanalysis data of the atmospheric wind field, satellite images of the spreading ash cloud, and the timing of the first ash fall at various distances from the volcano. Our results highlight the complex interaction between the co-PDC plume formed by the blast and the later vent-derived plumes from the Plinian phase, and emphasize, particularly, the ultimate impact of the blast on the characteristics of the fallout deposit.

2 Background

2.1 May 18, 1980 Mount St. Helens eruption: chronology

The chronology of the May 18, 1980 Mount St. Helens eruption has been well constrained by combining information from eye witness accounts (Criswell, 1987; Rosenbaum and Waitt, 1981), photographs and videos (Hoblitt, 2000; Moore and Rice, 1984; Voight, 1981), stratigraphic studies of the proximal deposits (Criswell, 1987; Hoblitt et al., 1981; Rowley et al., 1981; Sarna-Wojcicki et al., 1981; Waitt and Dzurisin, 1981), and radar and satellite observations of the ash plumes (Harris et al., 1981; Holasek and Self, 1995; Sparks et al., 1986). The eruption started at 08:32 in the morning (Pacific Daylight Time, PDT) with a catastrophic collapse of the northern flank of the volcano, which triggered the decompression of a shallow cryptodome and caused a blast directed towards the north that was accompanied by only a small (< 8 km a.s.l.) vertical plume at the vent (Glicken, 1990; Hoblitt, 2000; Moore and Rice, 1984). The blast generated a pyroclastic surge that covered an area of 572 km^2 to the north of the volcano in ~ 5 minutes (Druitt, 1992; Hoblitt et al., 1981; Moore and Sisson, 1981). It then uplifted abruptly at 10-15 km north of the vent and fed a huge convective column and plume that reached a maximum height of 30 km (Holasek and Self, 1995; Sparks et al., 1986). This blast-derived co-PDC plume (hereafter referred as co-blast plume) then dispersed towards the east, following the prevailing atmospheric winds (Fig.1; Holasek and Self, 1995).

The Plinian phase started soon after 09:00, and formed a weak to moderate vertical column at the vent that reportedly fluctuated around ~ 14 km height for the next three hours (Criswell,

1987; Harris et al., 1981). Curiously, this column is not evident in the GOES satellite images presented by Holasek and Self (1995). Eruptive activity during this time period generated only a few small PDCs that were emplaced on the upper flanks of the volcano (Carey et al., 1990; Criswell, 1987). Vigorous PDC activity started around 12:15, and was accompanied by vertically distributed earthquakes that recorded full connection of the deeper magma storage region to the surface (Scandone et al., 2007). Activity increased in intensity again between 15:00 and 16:00; resulting pumice flows on the north flank had a typical runout distance of 8 km (Rowley et al., 1981) and produced co-PDC plumes with an elongated source that extended 4-7 km north of the crater (Carey et al., 1990; Criswell, 1987). At 16:25, activity changed with the formation of a vertical column from the vent (Criswell, 1987) that reached a maximum height of 20 km before dropping to 6 km at 18:00 (Harris et al., 1981; Holasek and Self, 1995) as the eruption progressively waned (Fig. 1). The different plumes related to the Plinian activity all dispersed towards the east with the prevailing winds (Holasek and Self, 1995).

2.2 May 18, 1980 Mount St. Helens eruption: eruptive products

The eruptive products associated with the two phases of the eruption (blast and Plinian) are very distinctive. Juvenile material from the blast phase originated from the volcanic edifice, the cryptodome and the upper part of the conduit; as a result the juvenile component was degassed and microlite-rich (abundant plagioclase laths surrounded in places by poikilitic quartz (Cashman and Hoblitt, 2004; Hoblitt and Harmon, 1993). This material has a dacitic bulk composition with a highly evolved glass (>79 wt% SiO₂ measured in the most crystalline samples; (Blundy and Cashman, 2001)) and ~30% phenocrysts (Cashman and Taggart, 1983). Large volumes of variably fragmented material from the collapsed edifice were also ejected during this early phase (Druitt, 1992; Hoblitt and Harmon, 1993).

The Plinian phase, in contrast, primarily erupted highly vesicular and microlite-poor pumice and ash (Klug and Cashman, 1994) created during rapid ascent of magma from a storage system at intermediate depth (6 to 12 km; Scandone et al., 2007). The Plinian material had a dacitic bulk composition (63 wt% SiO₂), with a rhyolitic matrix glass (73 wt% SiO₂), and ~30% phenocrysts of amphibole, plagioclase, orthopyroxene, Fe-Ti oxides and rare clinopyroxene (Blundy and Cashman, 2001; Kuntz et al., 1981; Rutherford et al., 1985).

2.3 May 18, 1980 Mount St. Helens eruption: fallout deposit

The eruption produced a widespread fall deposit that extended across the states of Washington, Idaho, and Montana; the observed plume reached a maximum width of 200 km

and extended east for > 600 km before drifting south. The signature characteristic of the fall deposit is a secondary thickness and mass maximum that represents an over-thickening of ~ 30 mm at about 300 km from vent (Carey and Sigurdsson, 1982; Sarna-Wojcicki et al., 1981) (Fig. 2). Other curious features are the apparent northward displacement of the overall deposit relative to the vent, and a shift of the deposit axis from the ENE to E at a distance of ~ 200 km. Although these latter aspects of the fall deposit have never been discussed explicitly, all published attempts to model the deposit have required that the vent source be displaced tens of kilometres north of the mountain (Armienti et al., 1988; Fero et al., 2008; Folch et al., 2010).

The proximal fall deposit (20-60 km downwind) comprised a succession of 4 stratigraphic units that varied in grain size, juvenile type and lithic content (Criswell, 1987; Waitt and Dzurisin, 1981). Between 100 and 550 km from vent, only two layers were identified: a basal light-grey layer of fine ash up to 20 mm-thick, and an upper light-grey to tan layer of ash accounting for the rest of the thickness (Sarna-Wojcicki et al., 1981). This change in colour was interpreted to correspond to the colour change of the vent-derived plume at 12:15, which heralded the onset of ubiquitous PDC activity (Criswell, 1987; Sarna-Wojcicki et al., 1981). Yet, there is no documentation of a change in grain size or componentry associated with the change in colour in the deposit (Scheidegger et al., 1982). Beyond 550 km the deposit comprised only a single light-grey layer of fine ash (Sarna-Wojcicki et al., 1981).

Reconstructions of the total grain-size distribution (TGSD) reveal a deposit dominated by fine material (62 wt% of the deposit finer than 63 μm ; Durant et al., 2009), which is at the fine-grained extreme of TGSDs from well-documented explosive eruptions (Rust and Cashman, 2011). The grain-size distributions of 33 samples collected by USGS scientists during the days immediately following the eruption (Sarna-Wojcicki et al., 1981) were reanalysed by Durant et al. (2009) to re-measure, in particular, the smallest particles by laser diffraction methods. These samples come from five N-S transects that lie roughly perpendicular to the deposit axis and span from 160-630 km from the vent (Fig. 2). The data are remarkable not only in the extent to which individual samples depart from a log-normal shape (from bimodal in transects I and II to strongly skewed in transects IV and V) but also in the systematic change in grain-size distributions (GSD) shape from south (bimodal) to north (unimodal) along individual cross-wind transects (Fig. 3).

By averaging GSDs along each transect, Durant et al. (2009) demonstrated a global decrease of the mean diameter and an improvement of the sorting from transect I to V, consistent with size fractionation during ash dispersion and sedimentation from the plume (Bursik et al., 1992; Fisher, 1964; Sparks et al., 1992). By deconvolving individual GSDs, they

also identified a subpopulation (average mode 19 μm) that reaches a maximum proportion in transect III (320 km from the vent), which crosses the secondary maximum at Ritzville (Fig. 2). Consistent with earlier workers, they interpreted this subpopulation to record enhanced sedimentation by ash aggregation; they provided a new mechanism for aggregation, however, by suggesting that the timing of aggregation and fallout was controlled by hydrometeor formation as the particle-laden clouds descended through the ice melting point (at ~ 7 km altitude).

Hydrometeor formation, however, does not explain the extraordinary GSD variations along transects I, II and III (Fig. 3), which have not been addressed by any previous published studies. Transects I and II (at 160 and 250 km from vent, respectively) are particularly noteworthy in this regard, in that strongly bimodal GSDs towards the southern edge of the deposit transform dramatically to unimodal fine-grained distributions towards the north. Along transect III (through the maximum of the secondary thickness isopachs), GSDs from samples on the deposit edges (DAVIS1 and DAVIS 21) are quasi-lognormal, but GSDs across the secondary thickening axis also transition from pseudo-bimodal in the south to skewed unimodal in the north.

2.4 Deconvolving grain size of fallout deposits

Detailed studies of the grain size of fallout deposits have proved critical for deciphering both source fragmentation processes (Kaminski and Jaupart, 1998; Kueppers et al., 2006), and the dynamics of eruptive columns and dispersing plumes (Carey and Sparks, 1986; Koyaguchi, 1994). Analysis of individual GSDs within fall deposits provide further information on mechanisms of transport and pyroclast sedimentation, which alter the source grain-size distribution (Fisher, 1964; Pyle, 1989) by fractionation processes related to contrasting settling and flow behaviours of particles of different size and density.

The GSD of a pyroclastic deposit sample can be quantified by a probability density function, which is typically obtained by combining sieving ($> 90 \mu\text{m}$) and laser diffraction ($< 90 \mu\text{m}$) analyses (Eychenne et al., 2012). GSDs from fallout deposits are often complex, sometimes polymodal, and rarely follow an obvious mathematical function (Folk, 1966; Wohletz et al., 1989). In an attempt to reconcile the grain size of pyroclastic deposits with theory of fragmentation and transport, Wohletz et al. (1989) developed a model to deconvolve complex grain-size distributions (those that depart from a log-normal shape) into sets of lognormal subpopulations. Although initially developed to study sequential fragmentation during transport in PDCs, this approach has been broadly applied to tephra fall deposits (e.g.,

Durant and Rose, 2009; Durant et al., 2009; Taddeucci and Palladino, 2002). Importantly, this approach commonly generates more subpopulations than observed distribution modes. Additionally, there is little evidence that either primary or secondary fragmentation processes produce lognormal grain-size distributions (Rust and Cashman, 2011).

Eychenne et al. (2012) developed an alternative method to deconvolve polymodal distributions that limits the number of subpopulations to the number of identifiable modes in the raw distribution and does not assume that the subpopulations are necessarily lognormal. Instead, this method uses the versatile Weibull function, which can reproduce the shape of many probability density functions (e.g., lognormal, Gaussian and Rayleigh distributions).

3 Methodology

Here we re-examine two aspects of the Mount St. Helens fallout deposit. First, we reanalyse the grain size data of Durant et al. (2009) using methods developed by Eychenne et al. (2012). Second, we use scanning electron microscope images to identify individual components of samples collected both across the secondary thickness maximum and in one distal location. Together these analyses provide important new insight into the origin of the fallout deposit.

3.1 Processing of the grain size data

Nine of the 33 samples presented in Durant et al. (2009) show bimodal GSDs (in transects I and II; Figs. 2 and 3); in these samples, we separated the two distinct peaks by adapting the approach of Eychenne et al. (2012) described above. We extracted two subpopulations corresponding to the two modes of the raw bimodal GSDs using the fully automated DECOLOG 5.0 software (Bellotti et al., 2010; Caballero et al., 2014), which is based on an iterative algorithm for optimizing non-linear fitting of Weibull functions. Initial deconvolutions suggest that the bimodal distributions contain a fine subpopulation with very similar characteristics for all the samples, and a coarse subpopulation that varies in both shape and mode (Fig. 4). These two subpopulations will be hereafter referred to as CS_b and FS_b (for “Coarse and Fine Subpopulation from bimodal samples”, respectively). The probability density functions of the FS_bs were fitted by a Weibull function using a non-linear regression algorithm of the form:

$$f(x) = \frac{a}{b} \left(\frac{x-c}{b} \right)^{a-1} \exp\left[-\left(\frac{x-c}{b}\right)^a\right], \quad (1)$$

where x is the grain size in $\Phi = -\log_2(\text{grain size in mm})$, a is a constant controlling the shape of the distribution (from positively to negatively skewed to Gaussian), b is a constant controlling the spread of the distribution along the abscissa and c is a constant controlling the location of the distribution along the abscissa. The parameters obtained are summarized in Table S1 of the supporting information.

To assess whether the remaining 24 unimodal distributions could also be described by a constant fine subpopulation and a varying coarse subpopulation, we developed an algorithm to deconvolve the unimodal probability density functions into a fixed fine subpopulation (varying only within the range defined by the 9 FS_{bs}) and a coarse subpopulation of unknown shape. These two subpopulations will be hereafter referred to as CS_u and FS_u (for “Coarse and Fine Subpopulation from unimodal samples”, respectively). CS_u is modelled by a Weibull function described by equation (1) with the parameters a , b and c treated as unknowns. The amplitudes of CS_u and FS_u are also treated as unknowns. The algorithm allows the 5 unknowns to be adjusted via a graphical user interface until the best fit is obtained. The goodness of fit is measured using the coefficient of determination (R^2). In cases where the CS_u overlapped the FS_u by more than 70%, the deconvolution was not completed because the algorithm would not converge towards a unique solution. The mean, sorting, skewness and kurtosis of the raw and deconvolved distributions were calculated using the method of Folk and Ward (1957). The proportions of CS_u and FS_u within individual bulk distributions were also determined.

3.2 Componentry analyses

Samples DAVIS1, 10, 14, 17 and 21 from transect III, and sample BATE12 from transect V (Fig. 2), were also analysed for componentry. Polished sections of resin-mounted ash $< 90 \mu\text{m}$ in size ($>3.5 \Phi$) were prepared and imaged by backscattered electrons (BSE) on a Secondary Electron Microscope (Hitachi S-3500N, School of Earth Science, University of Bristol). Seven classes of particles were distinguished in the samples (Fig. 5A-D): (1) dense angular microlite-free glass (glass shards), (2) vesicular and microlite-poor glass, (3) phenocrysts (mostly broken) included in a vesicular microlite-poor glass, (4) variably vesicular microlite-rich glass characterised by contorted vesicle shapes, (5) moderately to highly oxidized and altered accidentals of various type (holocrystalline, basalt, basaltic andesite, etc.), (6) phenocrysts (mostly broken) included in a microlite-rich glass, and (7) phenocrysts (mostly broken) without adhering groundmass. The proportions within each sample of these seven components were determined by counting particles (≥ 430 per sample) defined by more than 2000 pixels ($\sim 16 \mu\text{m}$) on mosaics of BSE images. The componentry distributions in number %

were converted to weight % assuming a homogeneous distribution of all the components within each analysed grain size and a constant particle density for each individual component (Table 1). By using the mass per unit area measured in the field at the corresponding locations, we then determined the mass of each component per unit area (Table 2).

4 Results

In presenting the grain size data, we consider first the 9 bimodal samples, which are found only in the two crosswind transects that are closest to the vent (I and II at 160 and 250 km, respectively). We then consider the unimodal GSDs present along more northerly and more distal transects; of these 10 could be deconvolved (located in transects I, II and III) using the methodology described in Section 3.1, and 14 could not (DAVIS1, DAVIS17, DAVIS 21 located in transect III, and all the WOOD and BATE samples located in transects IV and V), because of an overlap between the two subpopulations of more than 70%. We use these data to examine downwind variations in grain size and component characteristics.

4.1 Variations of grain-size subpopulations within bimodal samples

The cumulative and probability density functions of CS_{bs} and FS_{bs} are presented in Figures 4 and 6, respectively. All deconvolved FS_{bs} have similar shapes and constant mean values of ~ 15 µm but the CS_{bs} vary strongly both along *and across* the deposit axis. CS_{bs} along transect I are substantially coarser than along transect II (mean values plotted in Fig. 7B and E), as expected for increasing distance from the vent. More unusual is the variation of the mean grain size of the CS_{bs} along the individual transects. Along transect I, the CS_b mean is largest at 35-40 km *south* of the deposit axis, and decreases in both directions away from this location (Fig. 7B). Along transect II, the CS_b mean is coarsest in the *southernmost* sample collected (100 km south of the deposit axis) and becomes slightly finer towards the north (Fig. 7E).

Modal proportions of different grain size fractions also vary both along and across the transport direction. CS_{bs} prevail over FS_{bs} along transect I (60-70 vs. 30-40 wt%, respectively; Fig. 7A), with a slight decrease in the CS_b proportion towards the north. The northernmost bimodal sample (DZ20-21) lies on the deposit axis and comprises 45 wt% coarse and 55 wt% fine subpopulations. In transect II, CS_b is preponderant within the most southern sample, and its proportion decreases abruptly towards the north (Fig. 7D). About 50 km south of the deposit axis along this transect, FS_b becomes the dominant grain size fraction.

4.2 Variations of the grain-size subpopulations within the unimodal samples

We successfully ($R^2 \geq 0.97$) deconvolved 10 unimodal samples in transects I, II, and III, although these samples generally show poor fits to the fine tail ($\leq 3\mu\text{m}$; Fig. 6). The resulting CS_u distributions complete the trend defined by variations in CS_{bs} . Along transect I, the two deconvolved unimodal samples are located to the north of the deposit axis, where the CS_u and FS_u proportions are similar to those measured in the northernmost bimodal sample and define a roughly constant trend from the axis to the northern edge of the deposit (Fig. 7A). Importantly, however, the CS_u distributions have smaller mean sizes than those of the CS_{bs} (Fig. 7B). The mass per unit area of both subpopulations reaches a maximum on the deposit axis, with both CS_u and FS_u mass decreasing towards the north (Fig. 7C).

The three CS_u measurements in transect II are also located within the northern half of the deposit and their proportion and mass per unit area are similar to those of transect I. Key characteristics include roughly constant CS and FS proportions to the north of the deposit axis (Fig. 7D), and mass peaks on, and ~ 20 km south of, the deposit axis for the coarse and fine subpopulations, respectively (Fig. 7F). Again the mean of the CS_u lies within the range of the CS_b mean values from transect I (Fig. 7E). In contrast, along transect III both the proportion and mean of the CS_u are constant (Fig. 7G and H). They are respectively lower (20 to 30 wt%) and smaller (mean value around $63\mu\text{m}$) than in transects I and II. Instead the fine subpopulations account for much of the mass (Fig. 7G), which reaches a maximum on the deposit axis for both CS_{us} and FS_{us} , consistent with the overall observation of a secondary thickness along this transect (Fig. 7I).

4.3 Downwind changes in GSD means

Figure 8 shows variations in the mean of the GSDs from all 33 fallout samples, as a function of the absolute distance from the vent. Where deconvolved, CS and FS means are plotted separately (bimodal data as filled circles, and unimodal data as open circles). First and foremost, the data show that while the mean of the coarse subpopulation decreases systematically with distance from vent, the mean of the fine subpopulation (FS_b) remains constant at $\sim 6\Phi$ ($16\mu\text{m}$). Two samples that depart from the CS trend are DZ20-25 and DZ20-32, which are located on the northern part of transect I. The non-deconvolved distributions have mean values close to those of the fine subpopulations, and occur at a distance at which the mean variation trends of the coarse and fine subpopulations converge.

4.4 Distribution of the ash components

Changes in the proportions of ash components (in both number % and mass %) across the secondary maximum (transect III) and in one of the most distal samples (transect V) are shown in Figure 5E. Along transect III, the abundance of vesicular microlite-poor particles decreases significantly towards the north (from ~30 to <5 wt%), while the proportion of accessory (non-juvenile) particles increases from ~16 to ~50 wt%. The other components do not show a clear spatial trend. The distal sample from transect V (BATE 12, collected at > 600 km) shows a similar componentry distribution to sample DAVIS 10 (near the secondary thickness maximum in transect III). It contains ~9 wt% of microlite-rich particles (componentry class 4; Fig. 5) and 41 wt% of microlite-poor juvenile particles (componentry classes 1 and 2). When corrected for mass the samples located within the zone of secondary thickening (DAVIS10 and DAVIS14) contain the same *absolute* amount of dense microlite-poor glass, while the absolute amount of vesicular microlite-poor particles decreases from the southern (DAVIS10) towards the northern sample (DAVIS14) (Table 2). Accessory and microlite-rich particles do not vary significantly.

5 Discussion

In the next sections we address the origin of both the different ash components and the bimodality of the grain-size distributions. We then examine the spatial variations of both the componentry and grain size in the context of reanalysis data of atmospheric wind fields and satellite images of the spreading ash cloud. Finally, we discuss different transport and sedimentation mechanisms likely to explain the depositional patterns highlighted by our results.

5.1 Origin and transport of the ash components

The different ash components identified in the fallout deposit (Fig. 5) can be related to the different phases of the eruption. The vesicular microlite-poor glass particles (Fig. 5C) were produced by rapid magma transfer from depth to the surface (with associated extensive vesiculation), consistent with the dynamics of the Plinian eruptive phase (Blundy and Cashman, 2001; Klug and Cashman, 1994). The dense glass particles (Fig. 5D) must come from the same vesicular magma, and represent the bubble-free melt interstices between larger bubbles. That these glass shards are not significantly smaller than the vesicular particles (Fig. 5C and D), indicates that the magma contained a wide range of bubble sizes, as inferred from textural studies of the Plinian products (Genareau et al., 2013; Klug and Cashman, 1994). Phenocrysts and phenocryst fragments included in vesicular microlite-poor glass can also be

attributed to the Plinian phase of the eruption. The microlite-rich particles, in contrast, are characteristic of the cryptodome material (Cashman and Hoblitt, 2004; Hoblitt and Harmon, 1993; Hoblitt et al., 1981; Muir et al., 2012). Also characteristic are the variable bubble contents of these samples and the contorted bubble shapes caused by expansion around groundmass microlites (Fig. 5B). Both phenocrysts and phenocryst fragments included in microlite-rich glass can also be attributed to the cryptodome, and thus the blast deposit. The early blast also incorporated abundant non-magmatic material from the collapsed mountain flank, which was observed in proximal blast deposits (Hoblitt and Harmon, 1993; Hoblitt et al., 1981). The generally altered nature of the accidental particles observed in the samples studied here (Fig. 5A and datasets S1 to S6 in supporting information) suggests material originating from the vicinity of a vent (either 1980 or earlier) or shallow subsurface, and erupted early, most probably during the blast phase. Phenocrysts with no observable adhering glass cannot be definitively identified as blast or Plinian, because phenocryst compositions in the juvenile material from these two eruption phases overlap (Berlo et al., 2007; Hoblitt and Harmon, 1993; Rutherford et al., 1985).

The blast-derived particles (cryptodome fragments, phenocrysts included in cryptodome-like matrix, and accidentals) were necessarily transported in the early co-blast plume (i.e. the co-PDC plume that uplifted from the surge produced by the blast in the early morning). The particles produced during the Plinian phase (dense and vesicular microlite-poor glassy particles, and phenocrysts in a microlite-poor matrix) could have been transported in the vent-derived plumes or in the co-PDC plumes generated by the afternoon pumice flows (hereafter referred as co-pumice flow plumes; Carey et al., 1990; Criswell, 1987). We further hypothesize that vesicular microlite-poor particles with sharp edges, thin walls and delicate shapes (cf. datasets S1 to S6 in supporting information) are unlikely to be preserved in a PDC where collisional and frictional processes cause abrasion and comminution of the flow material (Calder et al., 2000; Dufek and Manga, 2008; Manga et al., 2011). We thus infer that the vesicular microlite-poor particles were sourced from vertical activity at the vent, most likely during the late afternoon Plinian phase (Fig. 1).

5.2 *Origin of the grain-size subpopulations*

The observed downwind decrease in the mean grain size of the coarse subpopulation (Figs. 7B, E, H and 8A) is consistent with fractionation during dispersion of a laterally spreading volcanic plume containing a wide GSD (i.e. vent-derived plume), due to sedimentation governed by gravitational settling of individual particles in a turbulent to intermediate regime (Bonadonna et al., 1998; Bursik et al., 1992; Sparks et al., 1992). The crosswind variations of

the mean further suggest that the dispersal axis of the coarse material was displaced to the south of the (mass-defined) deposit axis (to ~30, 100 and 50 km south of transects I, II, and III depositional axes, respectively). The coarse subpopulations obtained from deconvolving the unimodal samples (by fixing the fine subpopulation) complete the trend observed in the bimodal samples, consistent with decreasing size of the coarse population and an assumed constant fine subpopulation at 12-16 μm , or 6 to 6.4 Φ (Fig.4 and 8A). From this perspective, the distributions that we were not able to deconvolve (DAVIS1, DAVIS17, DAVIS 21 and all the WOOD and BATE samples; Fig. 3) might also be a mixture of a constant fine and a slightly coarser subpopulation, which lies within the limiting 70% overlap in grain size that prevented deconvolution.

Similar patterns of bimodal grain-size distributions have been observed in fallout deposits from eruptions with sizes that range from intermediate (e.g., the 1997 Vulcanian explosions of Soufriere Hills, Montserrat, Bonadonna et al., 2002; and the August 2006 Tungurahua Subplinian eruption, Eychenne et al., 2012), to large (the June 1991 Pinatubo Plinian eruption, Wiesner et al., 2004) to very large and ignimbrite-forming eruptions (e.g., the Campanian eruption from the Phlegraean Fields, Engwell et al., 2014b; Sparks and Huang, 1980, and the Minoan eruption of Santorini, Sparks et al., 1984; Sparks and Huang, 1980). Fallout deposits from the latter two eruptions appear as coarse lower and fine upper units in proximal locations, as a massive ash layer with bimodal GSDs in medial locations, and as a massive unimodal ash layer in distal locations. In these examples, the mean grain size of the distal unimodal samples is identical to the fine peak of the medial bimodal ash layers (Engwell et al., 2014b; Sparks and Huang, 1980). Here the coarse subpopulation is inferred to derive from a Plinian eruptive plume formed at the vent, while the fine subpopulation is interpreted as ash settling from co-PDC plumes forming during emplacement of the (subsequent) ignimbrite flows (Engwell et al., 2014a; Engwell et al., 2014b; Sparks et al., 1984; Sparks and Huang, 1980). This interpretation is supported by both the fine grained nature of the ash layer and its relative enrichment in glassy material, which are characteristics of co-ignimbrite deposits observed on land (Sparks and Walker, 1977). The spatial pattern of 1991 Pinatubo fallout ash sampled in South China deep-sea cores has been interpreted the same way (Wiesner et al., 2004). Similarly, bimodal GSDs in the Tungurahua fallout deposit have been interpreted as mixed fallout and co-PDC ash, as demonstrated by correlating the enrichment trend in the deposit (determined from the proportion of the fine peak) to the zone of main PDC emplacement on the volcano's flank (Eychenne et al., 2012). At Montserrat, syn-eruptive PDC activity observed during the 1997 series of Vulcanian explosions has also been correlated with the occurrence of

bimodal grain- size distributions in the fallout deposit (Bonadonna et al., 2002). Critically, all examples show a constant fine grain-size mode at $\sim 5\text{--}6 \Phi$ regardless of the distance from vent; this characteristic may reflect the physical processes controlling the entrainment of particles from PDCs to co-PDC plumes, and/or the mechanisms governing their dispersion and sedimentation (see discussion below). The characteristics of the fine subpopulation retrieved from the bimodal samples of the Mount St. Helens fallout deposit are thus consistent with a co-PDC origin (either from the co-blast or co-pumice flow plumes).

5.3 A composite fallout deposit from multiple sources of ash

Both the grain size and componentry data indicate that at distances greater than 150 km from vent, the Mount St. Helens fallout deposit is composite, with material originating from both vent-derived (Plinian phase) and co-PDC (co-blast or co-pumice flow) plumes. The relative proportions of coarse and fine subpopulations, along with the crosswind variations of the coarse subpopulation mean, indicate that vent-derived particles are enriched to the south of the deposit axis, particularly in transects I and II (Fig. 7A-H). The componentry distributions in transect III also show both a relative and an absolute increase in Plinian phase particles towards the south, with a progressive increase in blast-derived particles towards the north (Fig. 5E and Table 2). Interestingly, componentry of distal sample BATE 12 shows that the distribution of components does not vary significantly along the deposit axis between 300 and >600 km from the vent, suggesting that blast- and Plinian-derived particles reached the same distance downwind, but were heterogeneously distributed across the plume dispersion axis.

Componentry alone cannot be used to distinguish particles transported in co-pumice flow plumes produced by the afternoon eruptive activity (Carey et al., 1990; Criswell, 1987; Fig. 1) from tephra produced by the vent-derived Plinian column. However, given the poor preservation potential of vesicular material containing thin bubble walls within a PDC (Dufek and Manga, 2008), co-pumice flow particles are likely to contribute primarily to the dense microlite-free glass, which does not show a crosswind trend. The correlation between the measured proportions of the fine subpopulation (FS) and the trends in componentry thus suggest that the co-PDC signature identifiable in the FS derives primarily from co-blast ash, although co-pumice flow ash might also contribute a modest amount. Overall, these findings demonstrate that the co-blast plume was an exceptionally efficient mechanism of fine ash formation and injection into the atmosphere, and produced a fallout deposit with an unusually

large amount of fine ash compared to other Subplinian and Plinian eruptions (e.g., Rust and Cashman, 2011).

5.4 Dispersion processes

The multiple sources of ash only partially explain the heterogeneity of the fallout deposit; additional explanations can be found in complex and changing dispersion processes with time. Here we constrain these processes by re-visiting original observations of plume transport and re-analysed wind data.

5.4.1 *Published observations of plume dispersion*

GOES visible images (at 8:50 and 9:20 PDT) show that less than 20 minutes after the eruption started, the circular umbrella cloud was ~ 80 km in diameter and centred north of the mountain (Holasek and Self, 1995; Pouget et al., 2013). It corresponds to the early co-blast plume, which uplifted from the entire area covered by the surge (≤ 15 km north of the vent; Sparks et al., 1986). According to classified satellite images (Sparks et al., 1986), the giant umbrella cloud reached a maximum height of >30 km in about 20 minutes, consistent with the maximum heights estimated from the GOES images at 8:50 and 9:20 PDT (Holasek and Self, 1995; Fig. 1). At 9:20, the top of the plume was still at 30 km and the umbrella region extended more than 100 km towards the north but *did not progress towards the south*; this asymmetry was a consequence of the strong northward momentum of the co-blast plume created by the high directional energy of the blast (Druitt, 1992; Esposti Ongaro et al., 2011; Hoblitt, 2000; Kieffer, 1981). The strong northward momentum of the plume is confirmed by photographs of the co-blast plume taken from the ground at 8:50, which show a very wide plume that increases rapidly in diameter with height and is bent to the north (Sparks et al., 1986). Critically, we suggest that this northward momentum explains the observed north-south asymmetry of the deposit isomass and isopach lines close to the vent (Carey and Sigurdsson, 1982; Sarna-Wojcicki et al., 1981; Fig. 2). Momentum-driven northward displacement of the plume may also explain the reported difficulties in replicating aspects of ash transport and deposition with advection-diffusion models, and the necessity of displacing the source tens of kilometers north of the actual source of the co-blast plume (Armienti et al., 1988; Folch et al., 2010).

Satellite images show that by 9:20, ash was spreading to the east and southeast at much lower altitudes than the 30 km plume top (see Plate 1a of Holasek and Self, 1995). By 9:50, this low altitude plume was strongly elongated to the east; at the same time the top of the plume (with a diameter of ~ 100 km), had detached from the vent and slowly progressed to the north. From 9:20 until at least 11:20, no plume is visible above the vent on the GOES images (Fig.

1), which might be a consequence of the large shadow cast by the high co-blast plume. Indeed, ground observers reported that a moderately high vertical column was emitted from the vent after 9:00 (Criswell, 1987). The maximum plume heights of 30 km measured by Holasek and Self (1995) during this time interval must thus represent the top of the co-blast plume, which subsided throughout the morning to an altitude of about 15 km. The vent-derived plume top was lower, and is probably what was measured by the meteorological radar located in Portland, Oregon (~ 80 km south-south east of the volcano), which reports a 14 km high plume during this time interval (Harris et al., 1981). This plume would also have dispersed towards the east.

5.4.2 Wind and ash dispersal directions

The wind speed and direction at the time of the eruption has been determined at different atmospheric levels using the trajectory model HYSPLIT, based on reanalysis meteorological data (Fig. 9). These data indicate that from 8:00 to 11:00 PDT, the wind was blowing towards the east at elevations of 10-20 km, with maximum velocities > 100 km/hr located near the tropopause (10-12 km). At higher elevations, the wind direction was toward the south-east at very low velocities; at the maximum altitude of the blast plume (30 km) there was almost no wind (Fig. 9A). Between 11:00 and 17:00 the atmospheric winds maintained an eastward direction and high velocities at altitudes of 10-15 km; the winds were dominantly southeasterly at elevations ≥ 18 km (Fig. 9A). After 17:00, the winds shifted to an east-south-east direction at elevations between 10 and 20 km, with a high velocity layer still located between 10 and 12 km (Fig. 9B).

The speed of the high velocity layer between 8:00 and 11:00 PDT is consistent with the average speed at which the plume dispersed to the east, as constrained by the satellite images (Fig. 1 and Table S2 in supporting information). These images show that both the co-blast and early-vent-derived material spread eastward in this high velocity layer at an altitude of ~10-12 km (Fig. 9A). The significant difference between the co-blast plume top height (30 km) and the height at which the material was preferentially dispersed (10-12 km) testifies to momentum-driven overshoot of the co-blast plume above the level of neutral buoyancy (Sparks et al., 1986), and to the control of wind conditions on the lateral injection height of ash into the atmosphere. Critically, the absence of wind towards the top of the co-blast plume limited dispersal at this altitude; for this reason, ash was dispersed eastward only after it settled into the high velocity layer near the tropopause. These complex proximal dispersion processes may explain why published advection-diffusion-sedimentation models have failed to reproduce the proximal area of the Mount St. Helens fallout deposit (Armienti et al., 1988; Fero et al., 2008),

and have produced an optimal fit of the medial to distal deposit trend by displacing the emission source well to the north of the vent (e.g., Folch et al., 2010).

Wind conditions between 11:00 and 17:00 (Fig. 1 and 9A) indicate that plumes (both vertical and co-pumice-flow) produced during the afternoon eruptive activity also dispersed towards the east in the 10-12 high velocity atmospheric layer. Co-pumice flow ash was thus distributed over approximately the same area as the co-blast ash, except that the co-pumice flow ash did not travel as far to the north because it lacked the northward lateral momentum. The shift in direction of the high velocity layer to the ESE after 17:00 (Fig. 9B) suggests that ash erupted during the late Plinian phase, which produced a plume of ≤ 20 km height, must have dispersed in a more southerly direction. This is consistent with both grain size and componentry data (Figs. 5, 7), which show that coarse mode of the Plinian ash was dispersed to the south of the main deposit axis.

5.5 Sedimentation processes

5.5.1 Sedimentation behaviours of the grain-size subpopulations

The contrasting grain-size patterns of ash deposits from vent-derived (Plinian) and co-PDC plumes reveal important differences in sedimentation processes. In proximal and medial areas, settling from vent-derived plumes, which, at these distances, are dominated by particles that are heterogeneous in size and coarser than about $> 125 \mu\text{m}$, can be explained by changes in individual particle settling velocities as grain size decreases with distance from vent (Bonadonna et al., 1998; Rose, 1993). Co-PDC plumes, in contrast, contain mostly fine material, irrespective of the type of PDCs they derive from, and deposit fine-grained material even close to source (e.g., 70% of co-ignimbrite particles $< 125 \mu\text{m}$; Sparks et al., 1973; Sparks and Walker, 1977). The widespread nature of co-PDC deposits, along with the constancy of their grain size with distance from source (e.g., Fig. 8), indicate that sedimentation cannot be controlled by individual settling processes of low Reynolds number particles (Bonadonna et al., 1998). Some other mechanisms must therefore govern the sedimentation of fine ash at the distances considered here. Distal (non-deconvolved) unimodal distributions also depict a constant mean grain size with distance from source, while componentry of sample BATE12 (Fig. 5) suggests that these samples still contain ash of mixed origin (i.e. co-blast and Plinian plumes). These unimodal distributions occur at a distance (both downwind and crosswind) where the vent-derived plume, depleted in coarser material, contains only particles of similar size to the co-blast plume (Fig. 8). Consequently, the constant GSD with distance (see also Durant et al., 2009) results from the specific sedimentation mechanisms governing the settling

of particles finer than a threshold size (around 63 μm), irrespective of the type of plume (vent-derived or co-PDC).

Profiles of mass per unit area (Fig. 7E, F and I) show that the sedimentation maxima of both the co-blast and vent-derived plumes (integrated throughout the whole eruption) are superimposed on the ground, despite some variations in their primary dispersal direction. Importantly, however, while the ash transported in the co-blast plume was injected into the atmosphere over a short time period (tens of minutes), vent-derived material was emitted at a variable discharge rate over the entire day (Fig. 1; Carey et al., 1990; Criswell, 1987). We suggest that the small grain size of the co-blast ash particles and the absence of wind at the top of the plume (Fig. 9A) allowed them to remain suspended for a long time, and, for this reason, that co-blast and vent-derived ash were most likely mixed *during transport*. Additionally, the superposition of the mass per unit area profiles (Fig. 7C, F and I) suggests that *sedimentation was enhanced where the two plumes overlapped*.

Additional information is provided by the recorded times of first ash fall (Fig. 1 and 2), which show that, consistent with the small size of the co-blast ash (Fig. 3), sedimentation was delayed towards the north relative to the south. From these data we conclude that early sedimentation of mixed bimodal ash in the south relative to the fine ash in the north could be related to the interaction of particles, whereby the large particles aided the sedimentation of the small particles (discussed below). In fact, time-series samples collected about 200 km downwind (locations WD and H on Fig. 2) show that the size distribution of ash falling during the periods of high accumulation rates was bimodal, and that more fine material was deposited near the dispersal axis than towards the southern edge of the deposit (Scheidegger et al., 1982). Although Scheidegger et al. (1982) make no mention of obvious ash aggregates or ash clusters falling at their sampling locations in the medial area (locations WD, PRD and H on Fig. 2) at any time during collection (from ~11:30 to 06:00 the next day), ash clusters were reported to fall: (1) at Pullman, WA, almost 400 km from the vent (Fig. 2), where GSDss are not bimodal (Fig. 3), and where the mean grain size has stabilized at $\sim 5.5 \phi$ (22 μm , Fig. 8; Sorem, 1982); and (2) at Kennewick, WA, located on the south edge of the fallout deposit, ~ 10 km south of sample DZ21-14 and ~ 230 km from the vent (Rose et al., 1982). It is important to note that these observations are among the first to have been made of falling ash aggregates, and that lack of aggregate observations in the medial area cannot strictly be interpreted as absence of falling aggregates.

Finally, mass per unit area profiles (Fig. 7A, F and I) show that the absolute amount of ash in the coarse (mostly vent-derived) subpopulation remains approximately constant downwind, even though the mean grain size decreases substantially (Fig. 8). In contrast, the absolute amount of ash in the uniformly fine (mostly co-blast) subpopulation increases sharply towards the area of maximum secondary thickening at about 300 km from vent (transect III, Fig. 7I). This observation suggests enhanced sedimentation of fine ash in this region (i.e. increased sedimentation rate), as suggested by previous workers (Carey and Sigurdsson, 1982; Durant et al., 2009).

5.5.2 Mechanisms for enhanced sedimentation

The process of enhanced sedimentation most commonly invoked to explain polymodal GSDs in distal fallout deposits and secondary thickness maxima is aggregation of fine particles to form coarser pellets or clusters; these aggregates are then deposited ‘prematurely’ due to their higher terminal velocity (e.g., Carey and Sigurdsson, 1982; Schumacher, 1994). More recent analysis further suggests that aggregate type changes from proximal to distal locations and aggregation is a temporally dynamic process (Brown et al., 2012). In proximal regions, where ash concentration in both vent-derived and co-PDC plumes is high, subspherical accretionary pellets (aggregates of concentric layers of particles sometimes organised around a larger particle) can be abundant (e.g., Bonadonna et al., 2002; Brazier et al., 1982; Brown et al., 2003). In medial to distal regions (> 100 km from source), only irregularly shaped loose ash clusters or particles coated by finer adhering grains have been observed (Bonadonna et al., 2011; Sorem, 1982). Although wet aggregation is a dominant process, the role of electrostatic forces (Gilbert and Lane, 1994), ice nucleation, and precipitation of secondary phases on particle surfaces is also significant, particularly in distal, weakly concentrated ash clouds (Brown et al., 2012; Gislason et al., 2011; Van Eaton et al., 2012).

Apart from the accretionary pellets recorded in the proximal blast deposits (e.g., Criswell, 1987; Druitt, 1992; Hoblitt et al., 1981), evidence of sedimentation of ash clusters in the medial and distal Mount St. Helens fallout deposit is rare (examples listed in section 5.5.1). An experimental study of electrostatic ash clustering performed on natural ash < 250 μm in size (Schumacher, 1994) roughly reproduced the downwind, *along axis*, grain size variations in the May 18, 1980 Mount St. Helens fallout deposit. This study reproduces grain-size distributions with a mode at 90 μm when single particles are coated with finer particles, while fine ash clustering produces wide distributions with a mode between 45 and 15 μm . Schumacher (1994) concluded that particle coating dominated in the proximal regions, while

ash clustering was the primary aggregation style in distal regions, and that both processes were of similar importance in the medial area, thus generating bimodal GSDs. This work does not explain the crosswind variations of grain size, however, and more importantly does not account for the prominent area of secondary thickness and mass observed in the isopach and isomass maps.

Numerical models of ash sedimentation that include an aggregation scheme were first constructed using empirical parameterizations (Armienti et al., 1988; Carey and Sigurdsson, 1982; Cornell et al., 1983), whereby particles smaller than a threshold size were assumed to form clusters of a given size and density, or to settle at a fixed terminal velocity. The early models applied to Mount St. Helens considered that all particles $< 63 \mu\text{m}$ aggregated (Armienti et al., 1988; Carey and Sigurdsson, 1982). While simulations of Carey and Sigurdsson (1982) reproduced the secondary maximum along the axis, they could not explain the proximal fine ash deposition. Armienti et al. (1988) also match the along-axis location of the secondary maximum but it lies 10-20 km south of the actual deposit, even though they moved the volcanic source 40 km north of the actual vent. A more complex model of wet aggregation was developed by Costa et al. (2010), who used a fractal law to describe the aggregation rate. Application of this model to Mount St. Helens also reproduces an area of secondary thickness (Folch et al., 2010), but the simulated deposits are much narrower than the observed deposit, are systematically displaced to the south, and the area of secondary maximum extends ~ 60 km closer to source than observed.

The presence of a secondary thickness and mass maximum at > 250 km from source implies a relatively abrupt change of sedimentation conditions. There is no obvious reason why aggregation rates would increase abruptly in this region given the absence of rain. Alternative models for enhanced sedimentation involve the bulk dynamics of the dispersing cloud rather than the behaviour of individual or clustered particles. They also account for the observation of pronounced mammatus at the base of the dispersing cloud above the medial area of the deposit (Durant et al., 2009; Schultz et al., 2006; Waitt et al., 2015). Durant et al. (2009) proposed a conceptual model for the formation of mammatus in volcanic clouds: ash particles in the cloud act as ice nuclei and ice hydrometeors form on ash particles, increasing mass and settling velocity. In this model, the high water vapour content of the Mount St. Helens cloud (derived from magmatic volatiles and entrainment of moist lower tropospheric air) drove abundant ice crystal growth. These hydrometeors accumulated at the cloud base and sublimated on contact with the ambient atmosphere, causing subsidence from the associated latent heat combined with gravitational loading. Moderately turbulent mammatus lobes form at some

critical threshold of ice particle number concentration and size (Schultz et al., 2006). As the cloud layer passed through the melting level at about 6.9 km altitude, the rate of aggregation increased rapidly in response to the sudden availability of liquid water, and a surge of loosely bound fine ash reached the ground, producing the distal mass accumulation maximum in the deposit.

Another model to explain the mammatus-like base of the cloud was proposed by Carazzo and Jellinek (2012; 2013) and relates to the development of “particle boundary layers” (PBLs) that retain small particles at the base of spreading plumes by a mechanism of diffusive convection. The development of PBLs is favoured by interaction of particles of different sizes and a high content of fine ash (Carazzo and Jellinek, 2013). These PBLs can also periodically detach from the cloud and settle en-masse due to gravitational instabilities (Carazzo and Jellinek, 2012). For example, gravitational instabilities observed close to the vent during the 2010 Eyjafjallajökull eruption (Iceland) channelled small and large particles to the ground at similar speeds (Manzella et al., 2015). Additionally, when the time of first ash fall from the Mount St. Helens plume (displayed in Fig. 2; Quinn, 1982) is compared with a numerical sedimentation model that did not include aggregation, not one but two cycles of delayed and accelerated ash fall with periods of 300 km can be seen (Fig. 4 of Hopkins and Bridgman, 1985). This cyclicity is consistent with a model of periodic en-masse release of particles. From this we suggest that the small absolute increase in the mass of the coarse subpopulation (which here has a mean size of 63 μm) in transect III, the significant increase in the fine subpopulation (with a mean size $\sim 15 \mu\text{m}$; Fig. 7I), and the asymmetry of individual GSDs across the region of secondary thickening (250 to 350 km from vent; Fig. 2) could all be explained by en-masse fallout from PBLs that were enhanced by the high fine ash content of the Mount St. Helens cloud. Additional aggregation and formation of hydrometeors within the boundary layers, as suggested by Durant et al. (2009), would have added to the fine ash fallout in the region of the secondary mass maximum.

Finally, as stated above, the observed crosswind variations of mass per unit area (Fig. 7) are at odds with simple aggregation processes and instead indicate enhanced sedimentation of small particles where coarser vent-derived material interacted with small (co-blast) particles during transport. Coating of coarser particles by fine ash is one possible mechanism; another is entrainment of small particles during (faster) sedimentation of larger particles. Particle coating could lead to higher absolute rates of sedimentation as the size difference between small and large particles decreases, because of an increase of the surface area available for

coating. This hypothesis provides an additional mechanism by which mass deposition may be enhanced in the area of secondary mass maximum, where grain-size subpopulations start to overlap (Fig. 6).

5.5.3 *Reappraisal of the mass eruptive budget*

Finally, the results of this study allow us to reassess the mass budget of the eruption. Sarna-Wojcicki et al. (1981) used planimetry to estimate the total mass of the fallout deposit from the isomass map shown in Figure 2 at $4.9 - 5.5 \times 10^{11}$ kg. We use our data to assess the contribution of different populations of ash grains (coarse and fine; blast and Plinian) in the analysed region (between transects I and III, i.e. ~ 150 to ~ 320 km from vent). By integrating both the mass per unit area across the plume axis (Fig. 7C, F and I) and the mass per unit length along the dispersal axis, we calculate a total mass of 1.2×10^{11} kg for this part of the deposit ($\sim 20\%$ of the total deposit calculated by Sarna-Wojcicki et al., 1981). Of this, the mass is approximately equally divided between the coarse and fine subpopulations (5.8 and 6.2×10^{10} kg, respectively). Considering that most of the fine subpopulation is likely to have a co-blast origin, these results indicate that co-blast ash accounts for at least 50% of the medial fallout deposit. The high contribution of co-blast material in the fallout deposit is also confirmed by componentry analyses performed by Scheidegger et al. (1982) on the time series samples collected at Wanapum Dam (Fig. 2), which show that microlite-rich particles dominated samples that account for 97 % of the total mass (that is, ash collected between 11:30 and 19:30 PDT). The large contribution of co-blast ash to the total deposit also explains the difficulties encountered in numerical models of ash dispersion and deposition (Armienti et al., 1988; Folch et al., 2010), all of which have thus far assumed that much of the distal mass was emitted from a Plinian column that persisted for the entire duration of the eruption.

Finally, the large contribution of co-PDC (and probably largely co-blast) ash explains the extremely small median grain size of the total grain-size distribution for this deposit (Rose and Durant, 2009; Rust and Cashman, 2011). Additionally, the componentry distributions show that at least 16 wt%, and plausibly as much as 52 wt%, of the ash in this part of the deposit is not juvenile (Fig. 5E). If this is representative of the componentry of the deposit as a whole, then the magma volume erupted on May 18 may have been overestimated by up to a factor of two.

6 Conclusion

The complex depositional pattern of the fallout deposit from the May 18, 1980 Mount St. Helens eruption results from the unusual eruptive dynamics, which strongly impacted the dispersion and sedimentation processes. The co-blast plume, in particular, was an exceptionally efficient mechanism for supplying fine ash to the atmosphere, and explains the remarkably high fine ash content of the fallout deposit from this eruption. Contributions to the sustained dispersing cloud from the two different main sources (co-blast and vent-derived plumes) created a pronounced crosswind asymmetry to both individual grain-size and componentry distributions, which could not have been unravelled without detailed analyses of these parameters in individual tephra samples. Fine ash is present throughout the deposit, but settled at an increased rate distally and reached the ground approximately 250-300 km from the volcano producing the area of secondary thickness and mass maximum. The sedimentation of fine ash was most likely controlled by a combination of enhanced settling mechanisms that include, but are not limited to, aggregation. Gravitational convective instabilities of cloud base particle boundary layers, hydrometeor formation and destruction due to the interaction of the ash cloud with different atmospheric layers, and entrainment of fine particles by larger ones, may also have played important roles. Together these processes ultimately determined the atmospheric lifetime of the fine ash.

Critical for the proximal sedimentation pattern was the horizontal (northward) momentum that drove the co-blast plume towards the north, eventually causing it to detach from the vent; at the same time, ash was progressively dispersed towards the east in the lower high-velocity winds at altitudes of 10-12 km. The early blast caused an overall displacement of the deposit towards the north as well as the north-south asymmetry of the isomass map, which has not yet been reproduced by advection-diffusion models. At distances beyond ~100 km from the volcano, ash was transported primarily within a high wind velocity layer at 10-12 km altitude, significantly lower than the maximum height of the blast clouds. This presents a cautionary note about the use of maximum plume height as a measure of the altitude of lateral dispersion of material in the atmosphere, or as a measure of the mass eruption rate. In this case the high vertical ascent rate was transitory.

Our findings also highlight the importance of considering individual phases of an eruption for the understanding of its dynamics, and the potentially contrasting behaviour of co-PDC and vent-derived plumes. In the case of the Mount St. Helens eruption, a critical analysis of the field data is essential to assess the impact of the complex eruption dynamics on the depositional

pattern of the fallout. This represents a critical step towards unraveling the combined effects of multiple eruptive processes and plume types on ensemble characteristics of eruptions, such as magmatic budgets, total grain-size distributions and fragmentation mechanisms.

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FIGURES

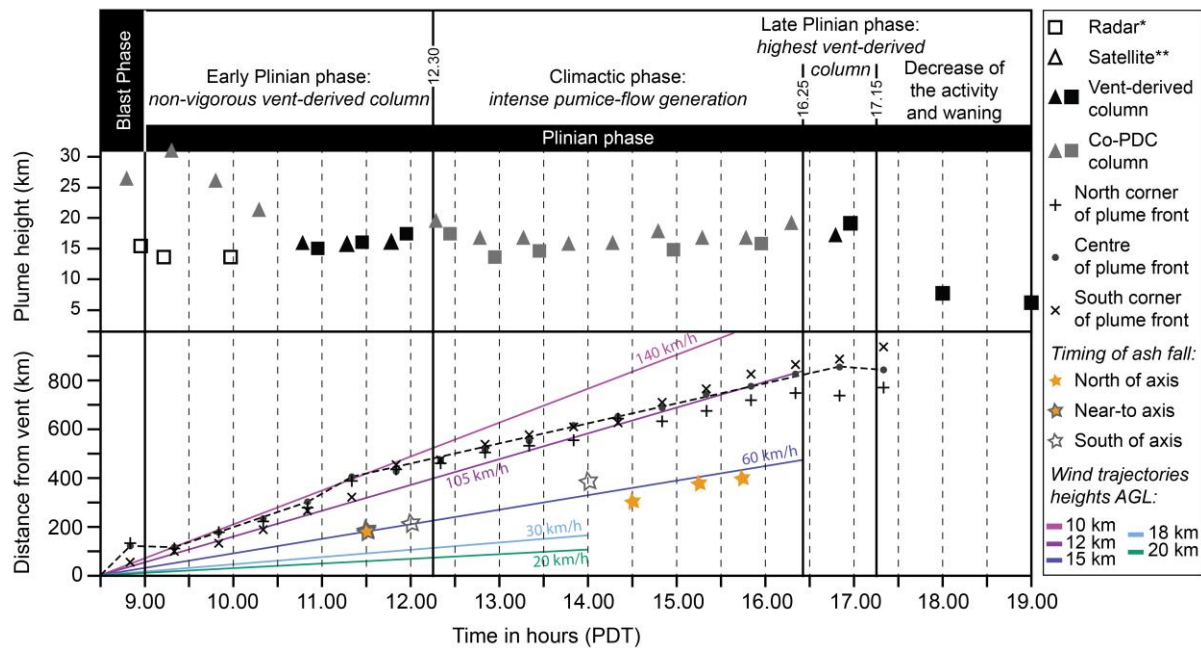


Fig. 1: Summary of the eruptive activity at the vent and of the plume dispersal throughout the duration of the eruption. The nature of the plumes (i.e. vent-derived vs. co-PDC) along with the timing and definition of the eruptive phases were inferred by cross-correlating plume height measurements, observations from the ground and interpretations from the literature (Carey et al., 1990; Criswell, 1987; Sparks et al., 1986). The extent of the dispersing plume towards the east was measured on the isochrone map built from the GOES visible images in Holasek and Self (1995). *Plume heights estimated using radar observations from Portland, Oregon (Harris et al., 1981). **Plume heights retrieved from GOES visible satellite images using the shadow method (Holasek and Self, 1995). The locations where timing of ash settling was observed are shown in Fig. 2, with references detailed in caption. The coloured straight lines in the bottom diagram represent eastward wind trajectories at different heights (calculated using HYSPLIT, see Fig. 9). The labels indicate the wind velocity.

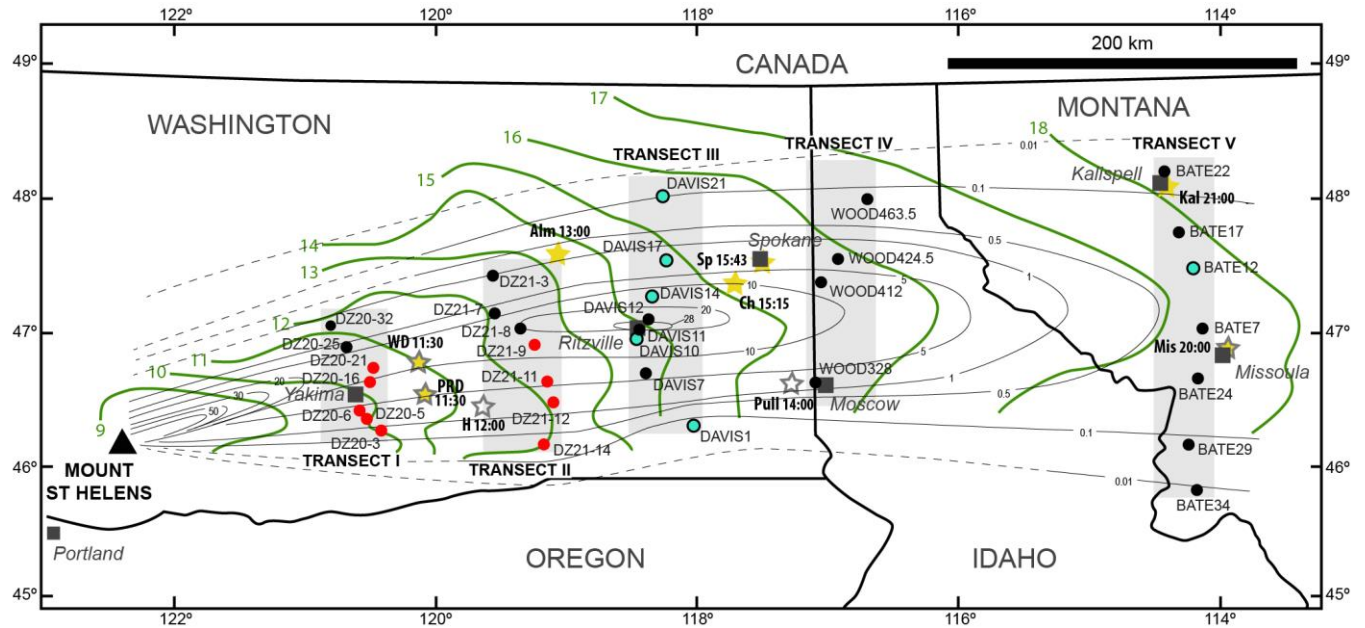


Fig. 2: Map of the May 18, 1980 Mount St. Helens fallout deposit picturing the isomass lines drawn by Sarna-Wojcicki et al. (1981) in kg/m² (thin grey contours). The dots represent the locations of samples considered in this study. Red dots are samples with bimodal grain-size distributions; dots with blue cores are the samples studied for componentry. Shaded zones define the five north-south transects of samples referred to in the text. Green lines are isochrones of first ash fall (Pacific Daylight Time) from Quinn (1982). Stars represent independent observations of first ashfall at specific locations, also reported in Fig. 1 (Pacific Daylight Time). Data from Scheidegger et al. (1982) (WD, PRD, H), Hooper et al. (1980) (Pull), Ikramuddin et al. (1982) (Ch), Stoiber et al. (1981) (Alm), Sarna-Wojcicki et al. (1981) (Sp, Mis, Kal).

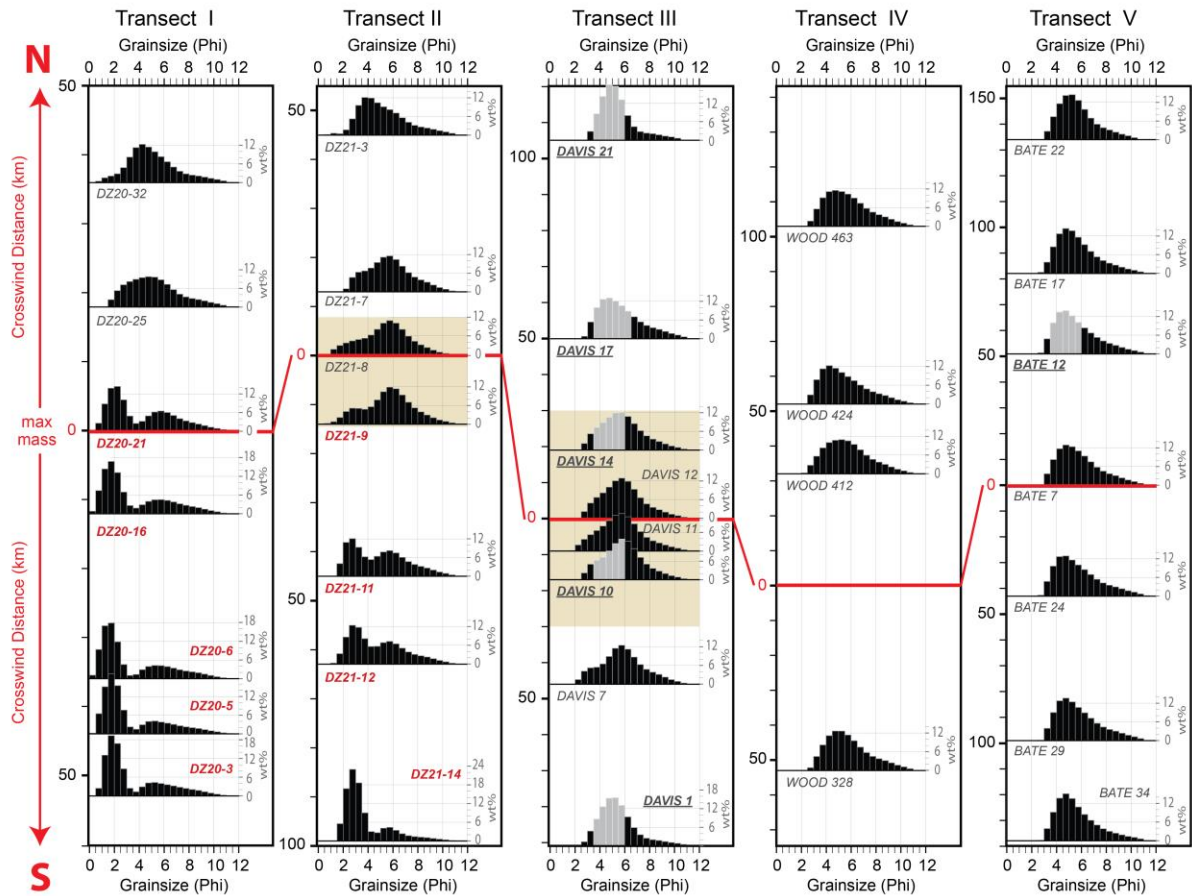


Fig. 3: GSDs of 33 samples from the May 18, 1980 Mount St Helens fallout deposit, as analysed by Durant et al. (2009). Distributions are organised by transects (Fig. 2) and plotted vertically at their projected distance (in km) from the deposit axis, represented by red horizontal line (location of maximum deposited mass). Note that the vertical distance axis has a different scale for each transect. Samples with bimodal grain-size distributions labelled in red; secondary thickness maximum highlighted by brown boxes. Samples analysed for componentry are underlined; approximate range of grain sizes analysed is represented by grey histogram bars. All GSDs plotted on same scale.

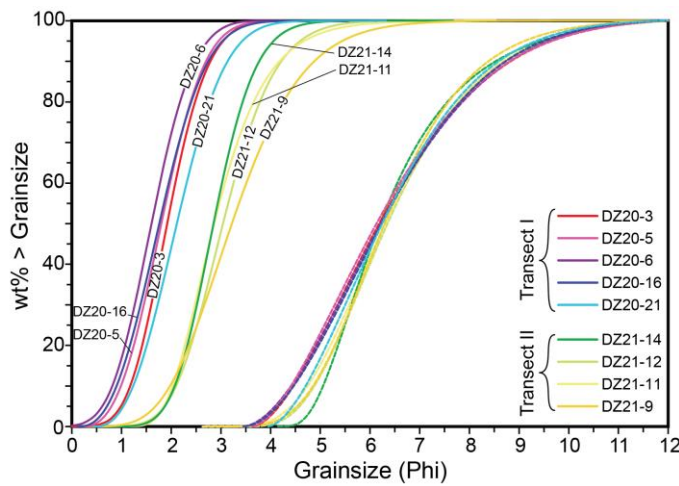


Fig. 4: Cumulative density functions of the deconvolved grain-size subpopulations (using DECOLOG 5.0) for the samples with bimodal GSDs (see Fig. 3). Solid lines represent the coarse subpopulations and dashed lines the fine subpopulations.

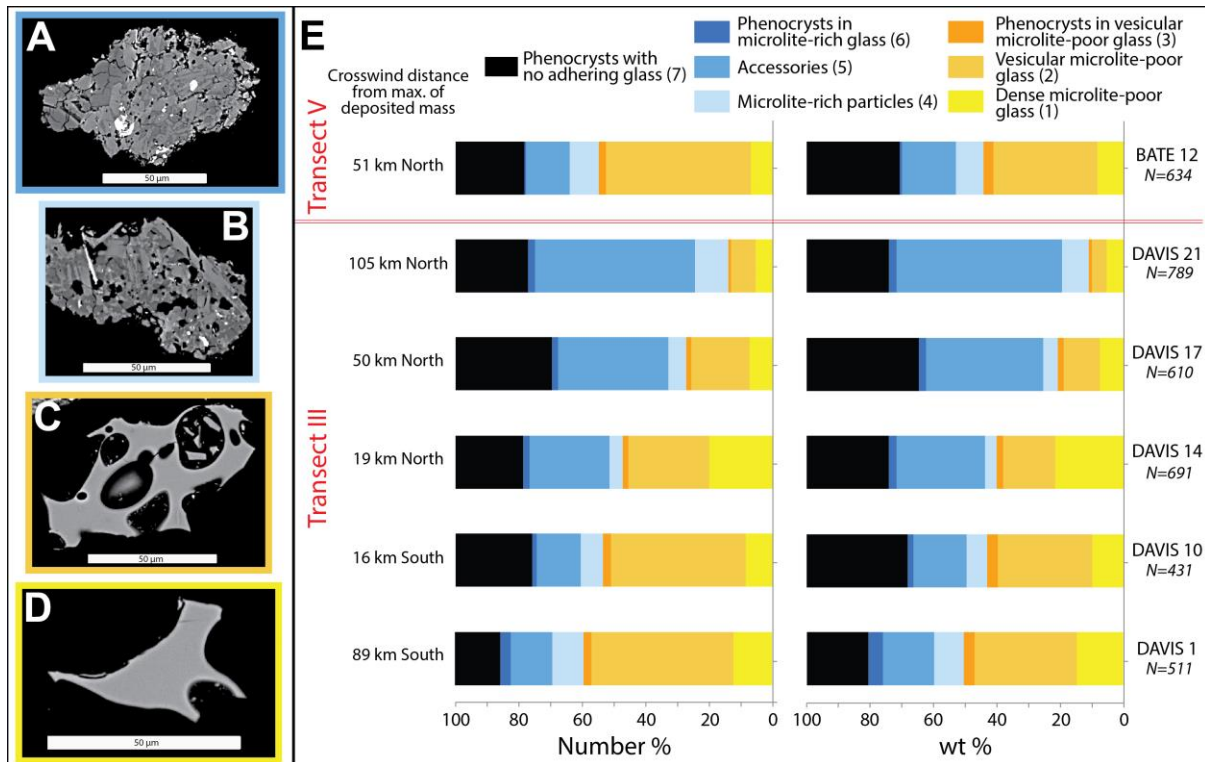


Fig. 5: Examples of particles from componentry classes: **A)** Accessory (5), **B)** Microlite-rich particles (4), **C)** Vesicular microlite-poor glass (2), and **D)** Dense glass (1). **E)** Componentry distributions in number and mass percent (see text for conversion protocol) within 6 samples from the fallout deposit. *N* is the number of particles counted. The number by the componentry class in the legend refers to component description in section 3.2.

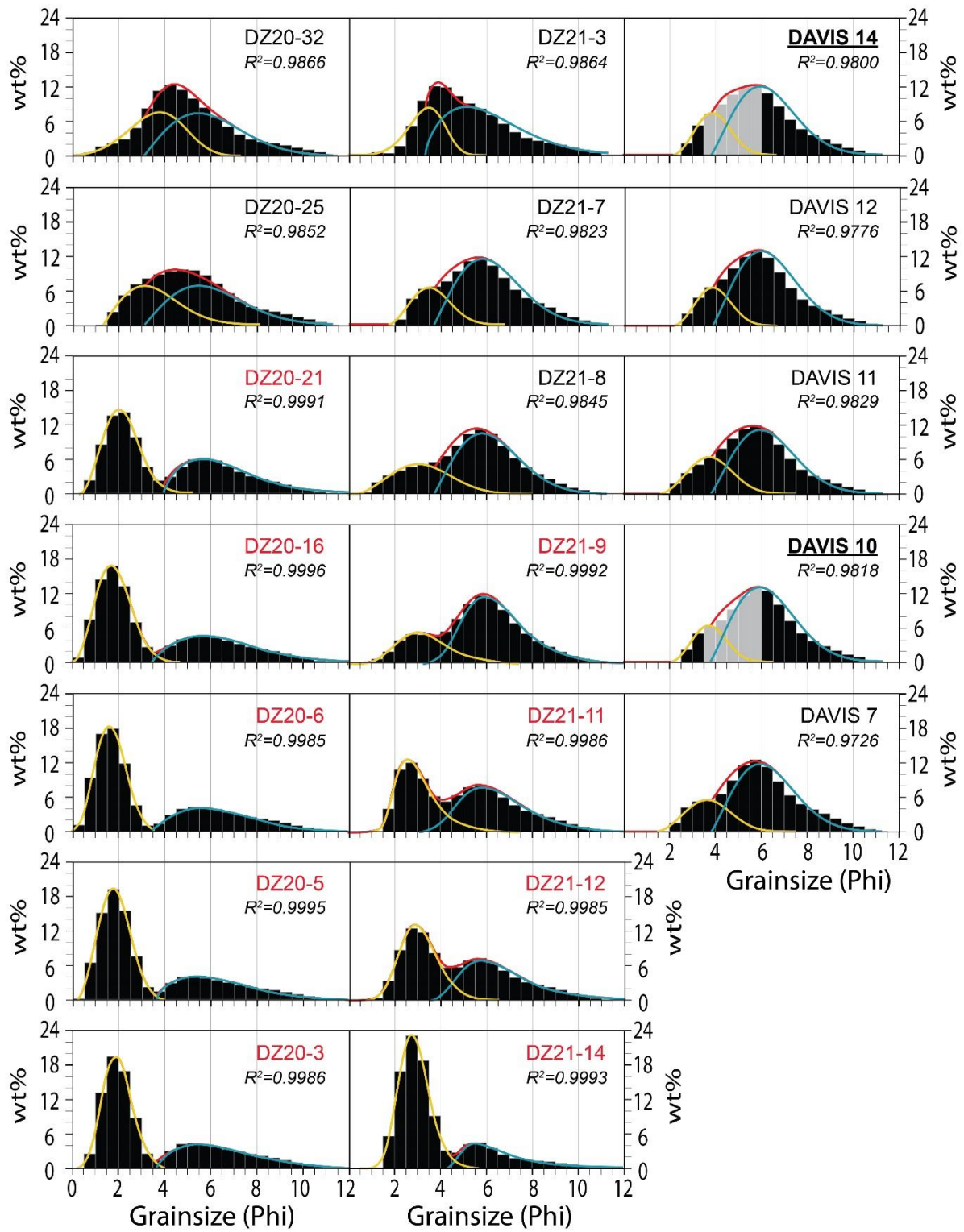


Fig. 6: Deconvolved GSDs, with fine subpopulation in blue, coarse in yellow and simulated bulk distribution in red. The bimodal GSDs are named in red and were deconvolved using DECOLOG 5.0. The other samples were deconvolved by fixing the shape of the fine subpopulation. R^2 (coefficient of determination) measures the fit quality produced by

deconvolution. Samples analysed for componentry are underlined, and approximate grain-size range represented by grey histogram bars.

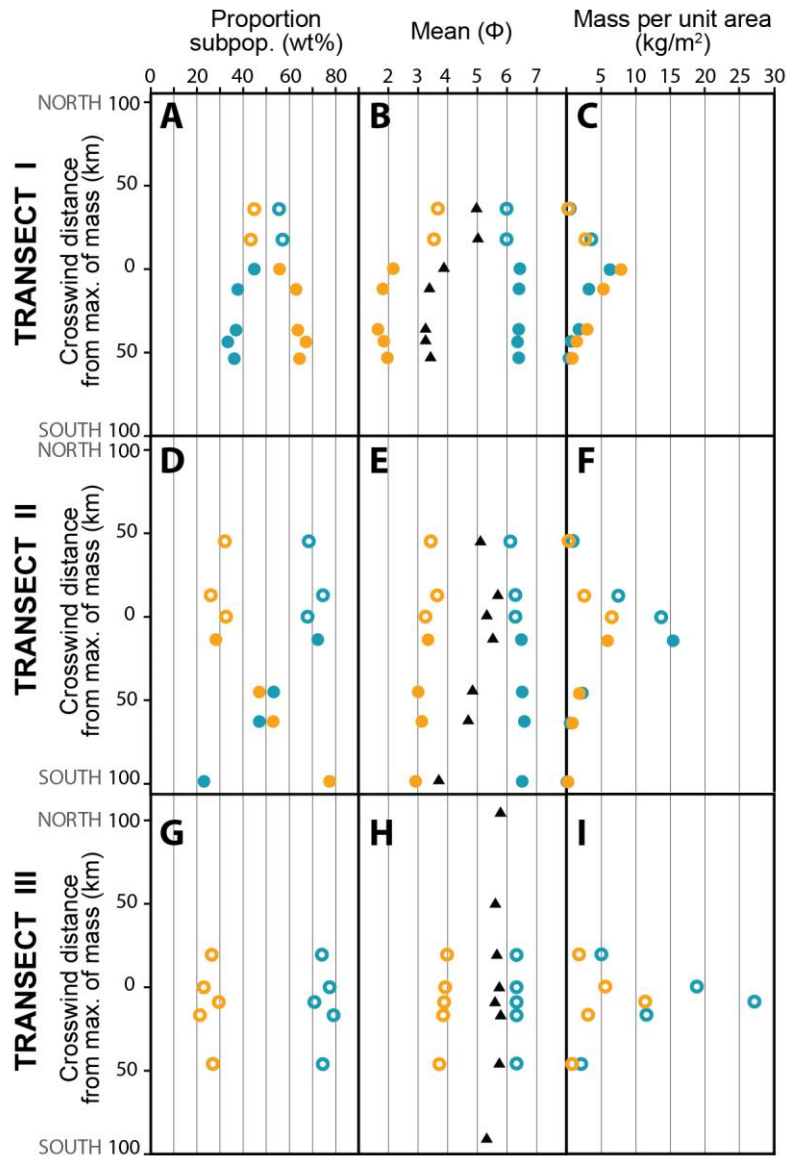


Fig. 7: Proportion (wt%), mean (Φ), and mass per unit area (kg/m^2) of the coarse (in orange) and fine (in blue) grain-size subpopulations obtained by deconvolution in transects I, II and III. Solid circles represent bimodal samples deconvolved using the fully automated DECOLOG 5.0, open circles represent unimodal samples deconvolved by fixing the fine subpopulation. Black triangles represent bulk distributions (before deconvolution).

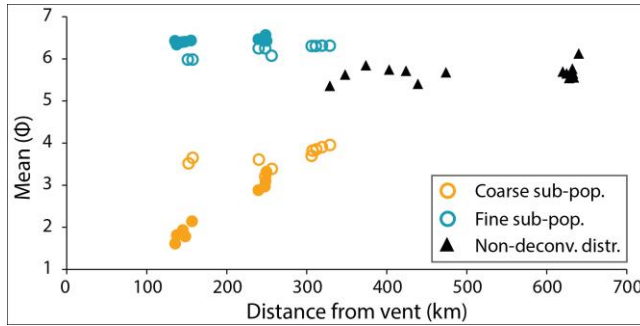


Fig. 8: Variation of mean (Φ) vs. absolute distance from vent (km) for coarse and fine grain-size subpopulations, and for non-deconvolved GSDs. Here solid circles represent the bimodal samples.

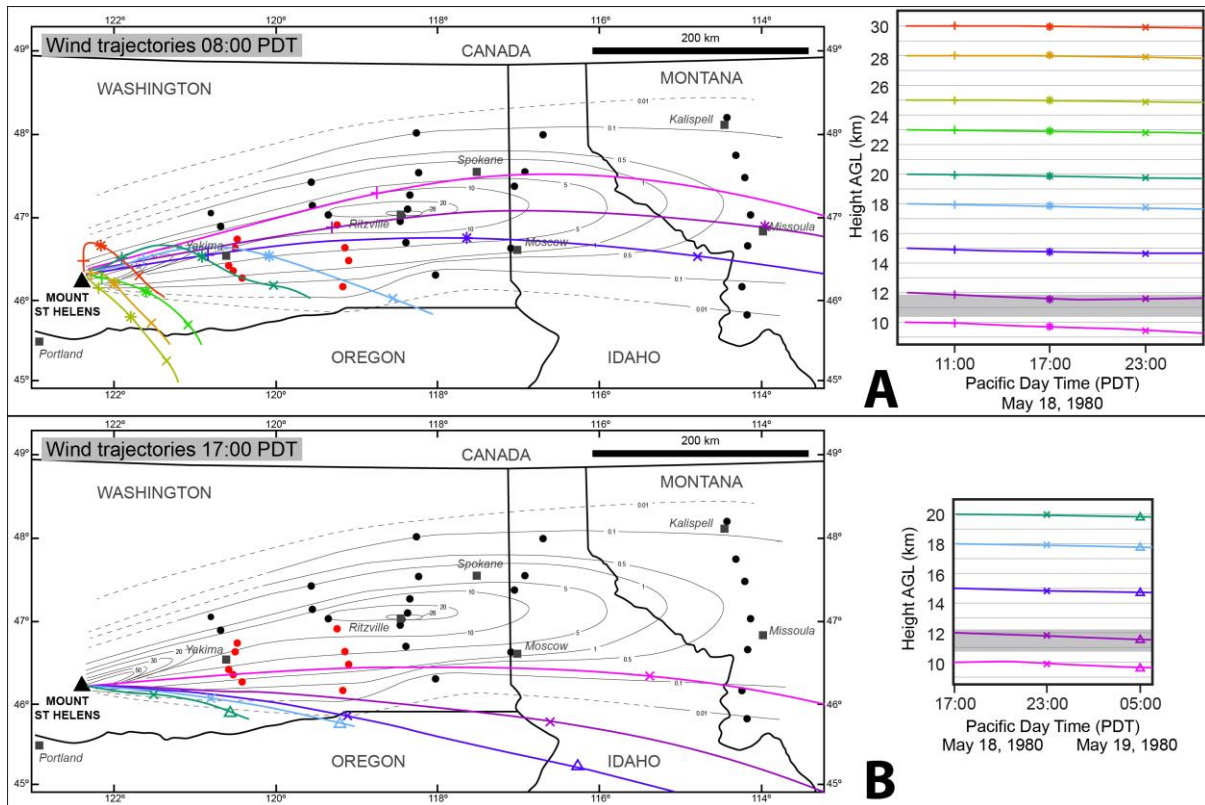


Fig. 9: Air parcel trajectories (analogous to wind trajectories) at different pressure levels for a source above Mount St. Helens volcano at 8.00 (A) and 17.00 PDT (B) on May 18, 1980, calculated using the HYSPLIT model (NOAA Air Resources Laboratory) and reanalysis meteorological data (NCEP/NCAR Global Reanalysis Data Archive). Right-hand diagrams represent corresponding heights (Above Ground Level) with time of each pressure-based trajectory plotted on maps. Each symbol type drawn along mapped trajectories represents a similar time and date (see right-hand diagrams). The grey shaded areas on the right-hand diagrams represent the approximate height of the tropopause (Holasek and Self, 1995).

TABLES

Componentry class	Mean particle density (g/cm ³)	References or assumptions
Dense microlite-free glass (1)	2.30	Cashman and McConnell (2005)
Vesicular microlite-free glass (2)	1.38	Mean vesicularity of 40%
Microlite rich particles (4)	1.83	Hoblitt and Harmon (1993)
Accessories (5)	2.37	Hoblitt and Harmon (1993)
Crystals (3, 6 and 7)	2.60	Cashman and McConnell (2005)

Table 1: Mean particle densities used to convert componentry distributions from number% to wt%. Numbers in first column refer to componentry categories in section 3.2.

<i>Sample</i>	<i>Mass per unit area (kg/m²)</i>							
	Total sample	Dense microlite-free glass	Vesicular microlite-free glass	Microlite-rich particles	Accessor.	Pheno. in microlite-poor glass	Pheno. in microlite-rich glass	Pheno. without adhering glass
DAVIS 1	1.8×10 ⁻²	2.6×10 ⁻³	5.7×10 ⁻³	1.7×10 ⁻³	2.9×10 ⁻³	6.1×10 ⁻⁴	8.0×10 ⁻⁴	3.4×10 ⁻³
DAVIS 10	14.7	1.5	4.4	9.5×10 ⁻¹	2.5	4.9×10 ⁻¹	2.7×10 ⁻¹	4.7
DAVIS 14	6.9	1.5	1.1	2.5×10 ⁻¹	1.9	1.4×10 ⁻¹	1.7×10 ⁻¹	1.8
DAVIS 17	9.1×10 ⁻¹	6.9×10 ⁻²	1.0×10 ⁻¹	4.2×10 ⁻²	3.4×10 ⁻¹	1.7×10 ⁻²	2.1×10 ⁻²	3.2×10 ⁻¹
DAVIS 21	6.0×10 ⁻³	3.3×10 ⁻⁴	2.8×10 ⁻⁴	5.0×10 ⁻⁴	3.1×10 ⁻³	6.0×10 ⁻⁵	1.5×10 ⁻⁴	1.5×10 ⁻³
BATE 12	1.4×10 ⁻¹	1.1×10 ⁻²	4.5×10 ⁻²	1.2×10 ⁻²	2.3×10 ⁻²	4.3×10 ⁻³	1.0×10 ⁻³	4.0×10 ⁻²

Table 2: Mass per unit area of different ash components at locations studied for componentry.

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